Use of precipitation and groundwater isotopes to interpret regional hydrology on a tropical volcanic island: Kilauea volcano area, Hawaii

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Abstract. Isotope tracer methods were used to determine flow paths, recharge areas, and relative age for groundwater in the Kilauea volcano area of the Island of Hawaii. A network of up to 66 precipitation collectors was emplaced in the study area and sampled twice yearly for a 3-year period. Stable isotopes in rainfall show three distinct isotopic gradients with elevation, which are correlated with trade wind, rain shadow, and high-elevation climatological patterns. Temporal variations in precipitation isotopes are controlled more by the frequency of storms than by seasonal temperature fluctuations. Results from this study suggest that (1) sampling network design must take into account areal variations in rainfall patterns on islands and in continental coastal areas and (2) isotope/elevation gradients on other tropical islands may be predictable on the basis of similar climatology. Groundwater was sampled yearly in coastal springs, wells, and a few high-elevation springs. Areal contrasts in groundwater stable isotopes and tritium indicate that the volcanic rift zones compartmentalize the regional groundwater system, isolating the groundwater south of Kilauea's summit and rift zones. Part of the Southwest Rift Zone appears to act as a conduit for water from higher elevation, but there is no evidence for downrift flow in the springs and shallow wells sampled in the lower East Rift Zone.

Introduction

Characteristic variations in deuterium and oxygen 18 content of precipitation due to differences in temperature, elevation, and distance from the coast have been used to infer recharge areas for groundwater, to indicate mixing, or to delineate different groundwater systems [e.g., *Gat*, 1971; *Fontes*, 1980]. Tritium originating from atmospheric bomb tests in the 1950s and 1960s enters groundwater in rainfall and has been used to estimate age of recent groundwaters. These isotope hydrology techniques are a particularly useful source of information about groundwater systems in areas with few wells.

Many volcanic islands have highly permeable surface rocks, resulting in ephemeral surface water and abundant spring discharge along the shoreline. Knowledge of groundwater flow patterns is generally limited; residents rely on groundwater or rain catchment systems for water supply, and the potential for groundwater pollution problems in permeable volcanic rock is high. Isotope hydrology methods seem particularly applicable in such areas, if there is enough isotopic variability to make them feasible. However, precipitation patterns on islands with appreciable topography can be complex, and precipitation isotope data from International Atomic Energy Agency (IAEA) tropical island stations commonly exhibit patterns different

from those found on other islands and continental stations [*Yurtsever and Gat*, 1981; *Dansgaard*, 1964].

In the southeastern part of the Island of Hawaii, comprising Kilauea volcano and the surrounding flanks of Mauna Loa volcano (Figure 1), perennial streams are absent despite an average rainfall of about 2000 mm/yr. Wells are generally located within 10 km of the coastline (below approximately 300 m in elevation), and there are no wells in a large part of the area. As a result, groundwater flow patterns in and around Kilauea volcano are not well known. The purpose of this study was twofold: (1) to use isotope techniques to interpret the hydrology of the Kilauea volcano area, particularly with regard to determining the regional hydrologic setting for Kilauea's geothermal systems, and (2) to determine precipitation isotope variation in an area with highly variable rainfall patterns, so the findings can be used in hydrologic studies in areas with similar climatology.

A network of as many as 66 precipitation collectors was emplaced in the 4200-km^2 study area and was sampled at approximately 6-month intervals from August 1991 to August 1994 (Figure 2). Groundwater wells, groundwater discharge along the coast (coastal springs and cracks), and high-elevation springs, totaling 53 sites (Figure 2), were sampled in August/September of 1991, 1992, and 1993. The precipitation and groundwater samples were analyzed for chemistry, deuterium (δD), and oxygen 18 ($\delta^{18}O$), and a subset of the groundwater samples were analyzed for tritium (^3H). A detailed report of
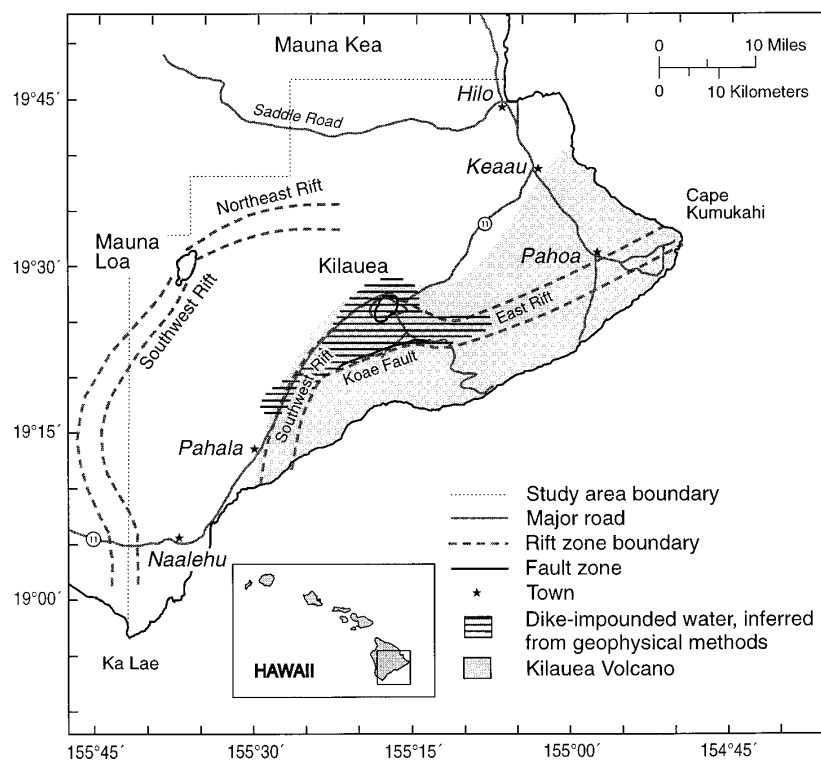
Figure 1. Map of the study area on the Island of Hawaii.

the study described here, including complete data tables, is given by *Scholl et al.* [1995].

Previous Isotope Studies on Islands

Environmental isotope data from other islands provide few coherent patterns to use as background for application to Hawaii. The worldwide IAEA data set yielded a meteoric water line of $\delta D = 6.17\ \delta^{18}O + 3.97$ for 15 tropical island stations [*Yurtsever and Gat*, 1981; *Dansgaard*, 1964], significantly different than the global meteoric water line of $\delta D = 8\ \delta^{18}O + 10$ [*Craig*, 1961]. Most published isotope studies on islands have focused on hydrologic interpretation of one catchment system. These studies generally used only a few precipitation or high-elevation spring samples to determine isotopic variations in recharge and provide no information on regional variations in isotopic content.

On Cheju Island, a volcanic island near Korea, *Davis et al.* [1970] used temporal variability in isotopic content of springs and wells as an index of the size of the groundwater reservoir. Rainfall records at Pohang, Korea (IAEA), were used to approximate rainfall on Cheju Island. At Pohang the most isotopically depleted precipitation occurs during the summer rainy season, whereas winter precipitation is isotopically enriched, a pattern opposite of that found worldwide [*Dansgaard*, 1964]. The isotopic enrichment of winter rainfall was attributed to its origin as first-stage condensation of moisture originating over the Sea of Japan. A study in the northern part of Taiwan also showed relatively depleted precipitation isotopic composition in summer and enriched isotopic composition in winter [*Liu*, 1984]. The isotopically depleted summer precipitation was attributed to typhoons, whereas isotopically enriched winter precipitation was believed to be derived from oceanic moisture close to Taiwan. For the Takaka River area in New Zealand, *Stewart and Williams* [1981] report the local meteoric water line

as $\delta D = 8\ \delta^{18}O + 13$, similar to the global meteoric water line. A 3-year record of monthly rainfall stable isotope content showed a seasonal pattern of enriched $\delta^{18}O$ values in summer and depleted $\delta^{18}O$ values in winter, similar to the worldwide pattern, except for a period of storms which produced isotopically depleted water during one late summer period. For the Piton de la Fournaise volcano on Reunion Island, *Nicolini et al.* [1989] found different meteoric water lines for two 6-month periods: $\delta D = 7.2\ \delta^{18}O + 6.0$ for a period without tropical storms and $\delta D = 8.0\ \delta^{18}O + 11.2$ for a period including a tropical storm. The first relation is similar to that found for tropical islands [*Dansgaard*, 1964], and the second relation is close to the global meteoric water line [*Craig*, 1961]. In the Martha Brae River basin, on the north coast of Jamaica, precipitation samples taken at three stations showed seasonal differences in δD gradients with elevation [*Ellins*, 1992].

Few previous studies report water isotope values on Hawaii. Published precipitation isotope data are available only for Hilo and the area upslope between Mauna Loa and Mauna Kea. *Friedman and Woodcock* [1957] measured δD in rainfall at four elevations up the Saddle Road (Figure 1) during a study of the trade wind–generated orographic rainfall. Deuterium values decreased with increasing elevation, but the small number of samples precluded any specific conclusions. *McMurtry et al.* [1977] collected stable isotope data for 20 springs and wells in the Kilauea area as part of a survey of potential geothermal sites. They found all the groundwater samples in the area to be depleted in D and ^{18}O relative to Hilo precipitation, but did not measure rainfall to obtain an elevation gradient. *Hsieh et al.* [1994] measured $\delta^{18}O$ in rainfall on the leeward side of Kohala Peninsula, the northern tip of the Island of Hawaii, and found values ranging from $-0.97‰$ to $-13.7‰$ along an elevation transect from 77 to 1250 m. Previous published tri-
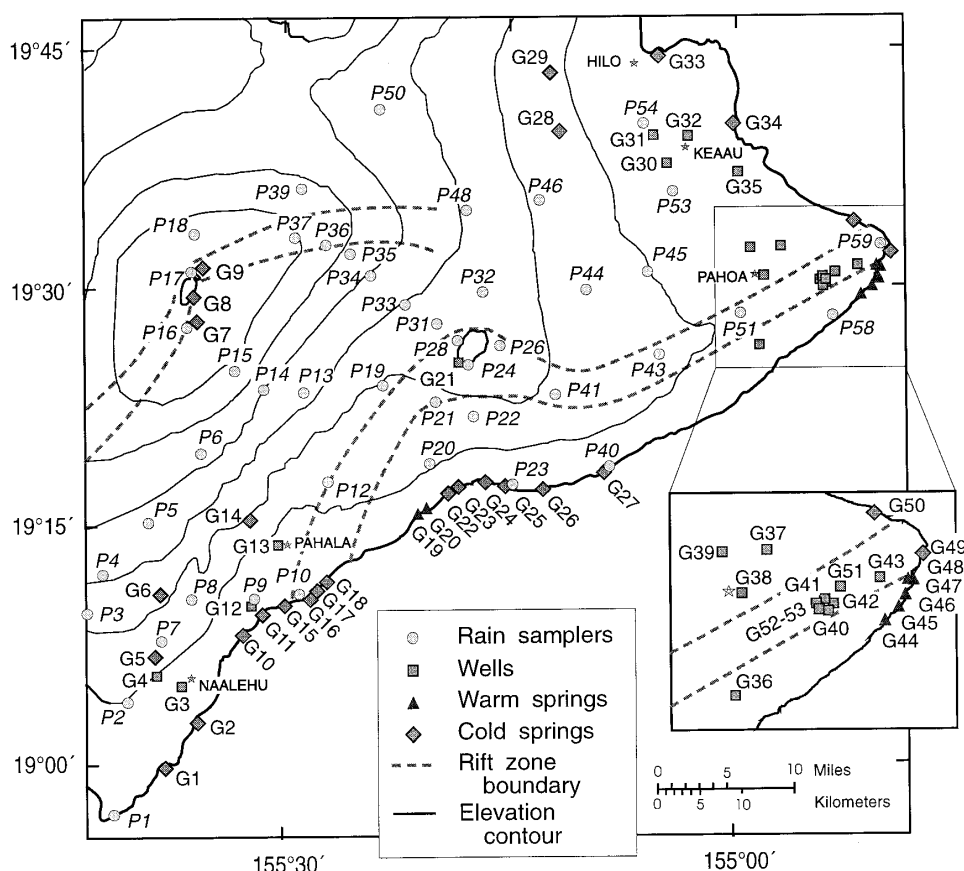
Figure 2. Locations of precipitation (P1–P59) and groundwater (G1–G53) sample sites. Elevation contour interval is 500 m.

tium measurements in the area are limited to the HGP-A well and some nearby water-supply wells near the lower East Rift Zone (ERZ) [*Kroopnick et al.*, 1978]. Stable isotope and tritium measurements in rainfall were published for stations in

Figure 3. Median annual rainfall in the study area, modified from *Giambelluca et al.* [1986]. Elevation contour interval is 500 m.

Hilo and on Midway, Wake and Johnston Islands from 1963 to 1984 (IAEA).

Climate and Rainfall on Hawaii

In the study area (Figure 1), mean annual temperatures range from 23.3°C at Hilo (sea level) to 7.0°C at Mauna Loa Observatory (3401 m). The climate of the Hawaiian Islands is dominated by trade winds from the east-northeast, which blow more than 90% of the time in summer and less frequently in winter. In winter, other weather systems (stationary highs, migratory highs, migratory low-pressure systems, and frontal systems) influence Hawaiian weather [*Schroeder*, 1993]. A temperature inversion, where air temperatures rise temporarily before continuing to decrease with increasing altitude in the atmosphere, occurs at approximately 2000 m elevation above sea level and is a persistent feature of the trade wind pattern. The inversion is a boundary above which the trade wind clouds evaporate.

The interaction of the trade winds with topography causes great areal variations in rainfall. Rainfall in the study area ranges from less than 500 to over 6000 mm/yr (Figure 3). The study area can be divided into three rainfall regimes; trade wind–dominated (south and west of Hilo, and Pahala-Naalehu), rain shadow (southwest of Kilauea summit), and high-elevation (above 2400 m on Mauna Loa). South and west of Hilo, the trade winds move moist air masses up the flank of Mauna Loa, causing the highest annual rainfall in the study area. The frequent rainfall on the southeast facing slope between Pahala and Naalehu is thought to be caused by a com-
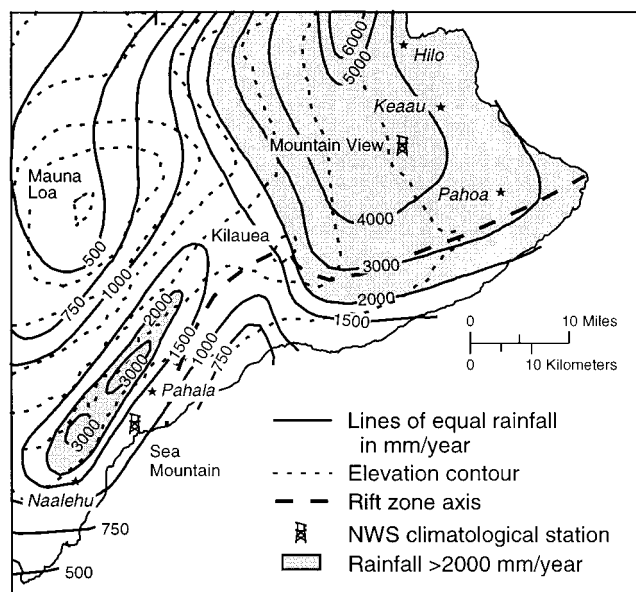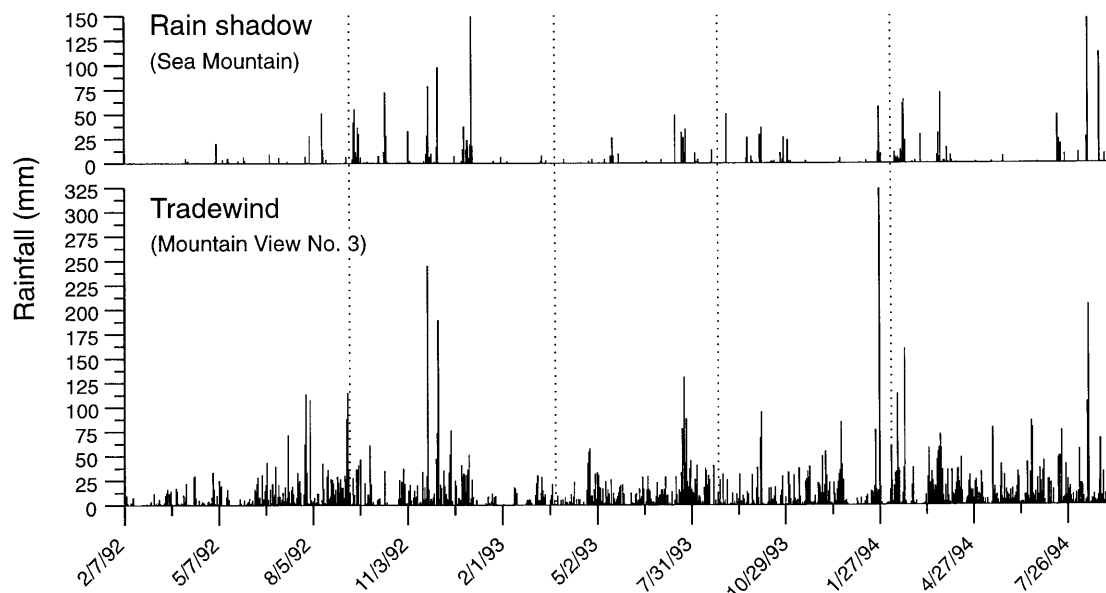
Figure 4. Daily rainfall in millimeters during the period of study, for representative rain shadow (Sea Mountain, index 51-8600-6) and trade wind (Mountain View 3, index 51-6546-6) stations [*NOAA*, 1992–1994]. Vertical dotted lines divide sample collection periods.

bination of trade winds and a thermally driven sea breeze/land breeze cycle [*Giambelluca and Sanderson*, 1993]. Trade winds flow downslope past the summit of Kilauea (1225 m elevation), causing a rain shadow on its southwest flank. This area receives rainfall from storm systems unrelated to trade winds or when trade winds are accentuated by a frontal system to the northeast of the islands (T. A. Schroeder, University of Hawaii, written communication, 1995). Winter frontal systems and "Kona storms" from the south are a major source of rainfall to the leeward sides of all the Hawaiian islands [*Schroeder*, 1993]. Kilauea summit itself has a sharp gradient in precipitation: the annual rainfall decreases by 1250 mm/yr from the northeast to the southwest side of Kilauea Crater [*Giambelluca and Sanderson*, 1993]. The upper slopes and summit of Mauna Loa (4145 m) are above the trade wind inversion. Rainfall above 2400 m elevation is infrequent, occurring only during storms.

Rainfall records for two weather stations with different precipitation regimes are shown in Figure 4 [*National Oceanic and Atmospheric Administration (NOAA)*, 1992–1994]. Mountain View station (location in Figure 3) is at 457 m elevation, midslope in the trade wind–affected area. There are very few dry periods; measurable rainfall occurs nearly every day. Sea Mountain station (location in Figure 3) is the station closest to Kilauea's rain shadow at which rainfall is measured daily. At this station, most rainfall occurs in discrete events.

Stable Isotope Composition of Rainfall

Sample Collection Methods

Precipitation collectors were placed to collect data in the trade wind–dominated, rain shadow, and high-elevation areas and sampled at 6-month intervals. In 1991–1993 a network of 66 collectors was emplaced (Figure 2). In 1993–1994 the network was reduced to 14 collectors. Collectors were designed to collect stable isotope samples; however, chemical analyses were also done on many of the samples to determine sulfate and chloride deposition patterns [*Scholl and Ingebritsen*, 1995].

Precipitation collectors were 13- or 19-L high-density polyethylene (HDPE) buckets with O-ring-sealed lids. Funnels of 5-, 8-, or 14-cm diameters were set in the lids, depending on the amount of rainfall expected at each site. At the end of the collection period the volume of rainwater was measured. Samples for isotopic analysis were passed through a coarse filter to remove oil and (or) particulate matter. The precipitation collectors were designed to minimize sample evaporation. For the first sample period (August 1991 through February 1992) most collectors had 5-cm funnels, and all contained a 3- to 4-mm layer of silicone oil. This design proved flawed; weight loss from indoor controls and an isotopic shift in the field control showed that there was evaporation through the layer of oil. The isotopic shift was enrichment along the meteoric water line, so that the evaporation would have been undetected in the absence of controls. Later field controls placed in shady to sunny locations showed weight losses of 0 to 33% and $\delta^{18}O$ enrichments of 0.5 to 6.2‰. Different collector designs were used for subsequent sample periods. At open-sky sites the funnel contained a plastic ball that floated when the funnel contained water and covered the opening at other times (M. L. Davisson, University of California, Davis, oral communication, 1991]. At forested sites the bucket contained a polypropylene bag and the funnel contained polyester fiber to exclude debris. A piece of plastic tubing attached to the funnel stem extended to the bottom of the bag, and both bag and bucket had 1-mm air outlet holes. The bag filled gradually, leaving little airspace to be occupied by vapor. For both these designs, evaporation measured by weight loss in controls did not exceed 0.04 g/d, and isotopic shift in field controls was undetectable. At a few sites, collectors with a thicker (8 mm) layer of oil were used. Controls with an 8-mm layer of oil showed enrichment in ^{18}O less than or equal to 0.6‰ over a 6-month period. Data from overfilled collectors were used at sites P13 and P15, where the data composed the only measurement available for these sites.

Apparent collector efficiency (rainfall in collector divided by rainfall in standard rain gauge) was estimated for sites where
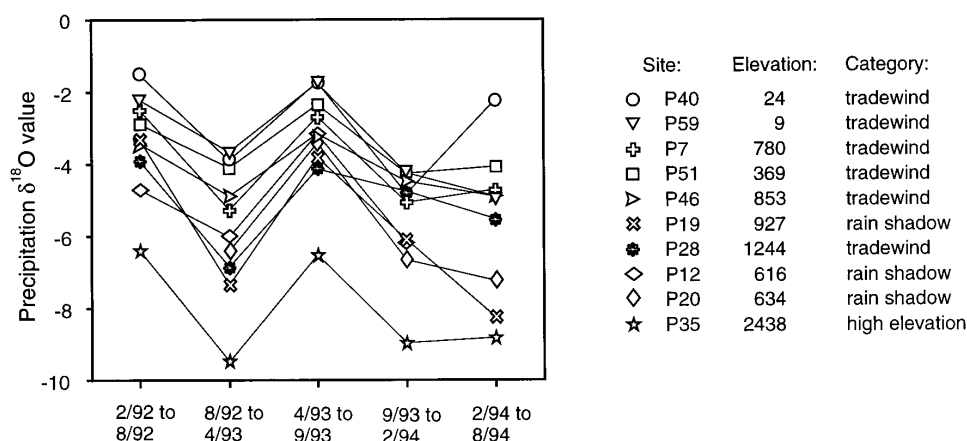
Figure 5. Temporal variation in $\delta^{18}O$ of precipitation at 10 sites where a sample was collected each time period.

we had a collector within 15 m of a standard gauge. Collector efficiencies ranged from 54 to 116%; most were within 85–110%. Precipitation in the study area was as much as 50% below normal during the first year of the study but near normal for the subsequent period.

Stable isotopes were analyzed at the U.S. Geological Survey, Menlo Park, California. Oxygen 18 was analyzed by the method of *Epstein and Mayeda* [1953], and deuterium was analyzed by the method of *Kendall and Coplen* [1985]. Values are normalized to standard light antarctic precipitation (SLAP) and reported relative to Vienna standard mean ocean water (VSMOW). Precision is $\pm 0.1‰$ for the $\delta^{18}O$ analysis and $\pm 1.5‰$ for the δD analysis. All stable isotope data for precipitation and groundwater are given by *Scholl et al.* [1995].

Calculation of Volume-Weighted Averages

Volume-weighted average δD and $\delta^{18}O$ values for precipitation samples were calculated on the basis of summer-winter pairs to obtain an annual average. If collectors were found to be nonfunctional or damaged, the sample was not analyzed for isotopes. For some sites there is only one summer-winter pair; for others, two summer values and one winter value; and so forth. Thus the volume-weighted average for each site was calculated in two steps: first, the volume-weighted average was found for each summer-winter pair, and then the volume-weighted average of all summer-winter averages was taken as final. Data from the first collection period were not used in the calculation of the average because the samples evaporated to a variable and unknown degree. The volume-weighted average samples for the study area define a local meteoric water line of $\delta D = 8.0\ \delta^{18}O + 12$, the same as that found by *McMurtry et al.* [1977].

Seasonal Variations

Data from the worldwide IAEA network have shown isotopic composition of rainfall in most areas to follow a seasonal cycle, with more depleted isotopic composition in winter and more enriched composition in summer, due to seasonal temperature variation [*Dansgaard*, 1964; *Yurtsever and Gat*, 1981]. The data in this study show a similar cycle for 1992 and 1993 (Figure 5), despite a relatively small seasonal variation in temperature (ranging from 2.8°C in Hilo, at sea level, to 3.8°C at Mauna Loa Observatory, at 3401 m). However, a seasonal cycle is not evident in the 1994 data, except at site P40 on the coast (Figure 5). Some samples from the 1994 spring-summer

period were even more isotopically depleted than the preceding fall-winter values.

Both the seasonal cycles in isotopic composition and the extremely depleted spring-summer values in 1994 can be explained to some extent by the rainfall record (Figure 4). During the 1992–1993 collection periods the number of storms was higher during the winter months at both the trade wind and rain shadow stations. (Storms are defined here as one or more consecutive days of rainfall greater than 50 mm). However, during the summer of 1994 there were more storms than in the previous summer periods. Trade wind rains are generated no higher than the inversion at 1800–2400 m elevation, where the minimum temperature is about 11°C. Storms extend much higher into the atmosphere, with condensation at 4000–6000 m (T. A. Schroeder, University of Hawaii, written communication, 1995). Lower condensation temperatures cause the water isotopes to undergo a larger degree of fractionation. Also, in contrast to the trade wind–generated orographic rainfall, which is the first condensation from ocean-derived moisture, large storm systems may have been raining for some time before they reach the island. The lower temperatures, combined with the likelihood that storm systems are raining before they reach land, would lead to much more depleted isotopic composition for storm precipitation. As noted previously, depleted isotopic composition in storm precipitation has been reported for northern Taiwan [*Liu*, 1984] and for hurricane and tropical cyclone events in the Caribbean Sea and Indian Ocean [*Nicolini et al.*, 1989]. We collected one isotope sample directly from a tropical storm (at site G45; Figure 2) that had isotopic composition of $\delta^{18}O = -5‰$, $\delta D = -32‰$, whereas the longer-term volume-weighted average for that area is $-3.2‰$, $-15‰$ (site P58; Figure 2).

Although our data only cover a 2½-year period, they suggest that the proportion of storm precipitation versus trade wind–driven or thermally driven orographic precipitation controls seasonal variations in the isotopic composition of precipitation in Hawaii. Higher frequency of storms is usually, but not always, correlated with the winter months. An instance of isotopically depleted summer rains similar to that seen on Hawaii in 1994 was seen in the 3-year record from New Zealand [*Stewart and Williams*, 1981]. Isotopically depleted summer precipitation appears to be more common in areas with monsoon climates, such as Pohang, Korea [*Davis et al.*, 1970], or Taiwan [*Liu*, 1984]. Seasonal temperature differences must also influ-
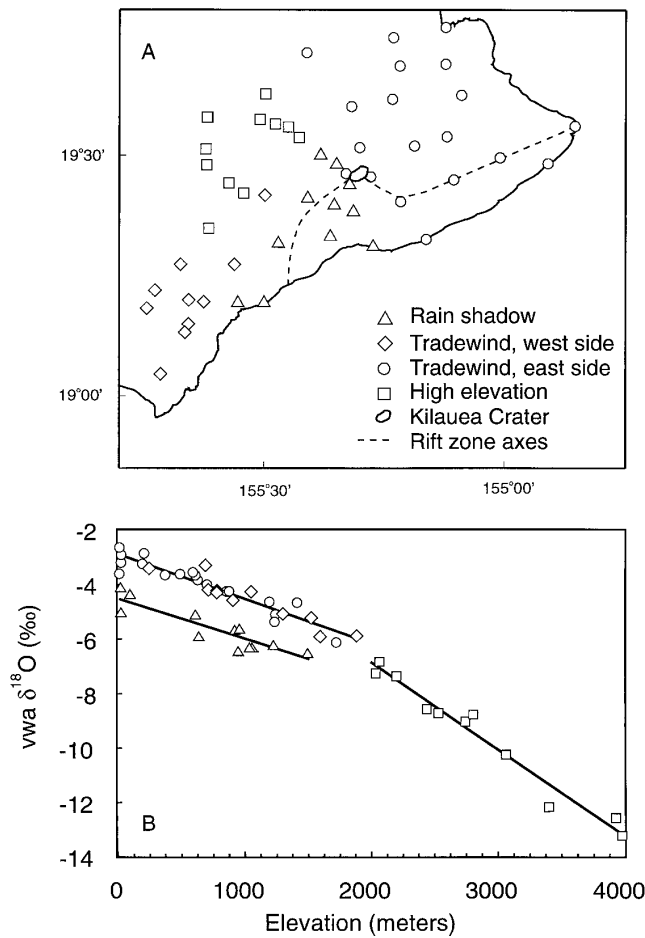
Figure 6. (a) Map of study area showing sites used in equations (1)–(3). Symbols denote climate classification of site as defined by isotope/elevation relations shown in Figure 6b. (b) Relation between volume-weighted average (vwa) $\delta^{18}O$ values and elevation in the study area.

ence precipitation isotopes on Hawaii, as there are seasonal variations in areas that receive only storm rainfall (Figure 5).

Stable Isotope/Elevation Relations

Volume-weighted annual average rainfall $\delta^{18}O$ values were plotted against elevation of the sample site, and regression lines were determined for the trade wind, rain shadow, and high-elevation areas (Figure 6). Hilo rainfall (IAEA), high-level spring, and stream data were included with the volume-weighted average rainfall data for the regressions. The relation between precipitation isotope values and elevation in meters (h) defines three distinct linear trends that are correlated with the local climatological patterns:

Trade wind areas ($n = 29$):

$$\delta^{18}O = -0.00164(h) - 2.85 \qquad r^2 = .90 \qquad (1a)$$

$$\delta D = -0.0123(h) - 11.2 \qquad r^2 = .80 \qquad (1b)$$

Kilauea's rain shadow ($n = 11$):

$$\delta^{18}O = -0.00148(h) - 4.44 \qquad r^2 = .80 \qquad (2a)$$

$$\delta D = -0.00978(h) - 26.7 \qquad r^2 = .75 \qquad (2b)$$

High-elevation area ($n = 11$):

$$\delta^{18}O = -0.00319(h) - 0.45 \qquad r^2 = .97 \qquad (3a)$$

$$\delta D = -0.0259(h) + 9.3 \qquad r^2 = .96 \qquad (3b)$$

In the "east side" and "west side" trade wind rainfall areas (Figure 6) the $\delta^{18}O$-elevation relation is the same. In Kilauea's rain shadow, rainfall data fall along a trend that has a slope similar to the trend for the trade wind areas but has an intercept that is about 1.6‰ more depleted in ^{18}O. The high-elevation data follow an elevation trend with a slope distinctly different from either the trade wind or rain shadow data. The trade wind and rain shadow gradients for Hawaii fall within the range of elevation/isotope gradients found for other islands (Table 1). The high-elevation gradients for Hawaii fall outside the range, likely because most islands do not have such high elevations.

Figure 7 shows the same volume-weighted average $\delta^{18}O$ data plotted against elevation, along with a Rayleigh condensation curve (see *Dansgaard* [1964] for processes controlling isotopic composition of rainfall) and the range of $\delta^{18}O$ composition for condensate in equilibrium with oceanic vapor at temperatures between the land surface and the top of the inversion layer. The relation between precipitation isotopes and elevation defined by most of the low-elevation data can be explained by mixing between depleted vapor from which some condensation has occurred and "fresh" oceanic vapor and/or by mixing depleted vapor with vapor returned to the atmosphere by transpiration. The trade wind rainfall data from the lowest elevations fall in the shaded range on Figure 7, indicating that orographic lifting and condensation of water vapor originating over the ocean is the predominant mechanism of rainfall in both trade wind areas.

The land-surface slope is different in the two trade wind areas and, as noted previously, the "west side" (Pahala-Naalehu area) rainfall is attributed to a thermally driven sea breeze cycle as well as to trade winds. The similarity of rainfall isotopic composition between the geographically and topographically distinct "east side" (Hilo-Pahoa) and "west side" (Pahala-Naalehu) areas (Figure 6) suggests that the same isotope/elevation relation may apply to climatologically similar areas on other islands, especially the other Hawaiian Islands, where the starting composition of the water vapor and the temperature gradient with elevation would presumably be the same.

The more depleted isotopic content of rainfall in the rain shadow of Kilauea volcano likely is due to the fact that the area receives only storm rainfall. In many continental areas leeward of mountain ranges there is little suggestion of an isotopic gradient [e.g., *Smith et al.*, 1979; *Ingraham and Taylor*, 1991; *Ingebritsen et al.*, 1994; *Adams et al.*, 1995]; however, there is a reasonably well defined gradient in Kilauea's rain shadow. In windward areas on Hawaii there is a consistent orographic lifting effect, with upslope wind flow accompanying rainfall. This may be the case in the rain shadow as well, where wind-direction data indicate that 41–63% of large rain events during the study period were associated with storms that came onto land from a southerly direction [*Scholl et al.*, 1995].

At high elevations on the summit of Mauna Loa the rate of decrease in δ value with elevation is greater than in the lower-elevation areas (Figures 6 and 7). We assume that the only source of precipitation is storms, because the area is above the trade wind inversion. *Smith et al.* [1979] presented a theoretical

Table 1. Gradients of Isotopic Composition With Elevation From Islands

| | Samples | | Isotope/Elevation Gradient, |
Location	Number	Type	per 100 m
Negros Island, Philippines*	4	precipitation	$\delta^{18}O = -0.19\%o$
			$\delta D = -1.14\%o$
Greater Tongonan geothermal field, Philippines†	3	precipitation	$\delta^{18}O = -0.07\%o$
			$\delta D = -0.25\%o$
Tatun Shan, Taiwan‡	32	spring, stream	$\delta^{18}O = -0.15\%o$
Mt. Yatsugatake, Japan§	10	precipitation	$\delta D = -1.69\%o$
New Zealand (no location given)‖	···	···	$\delta^{18}O = -0.23\%o$
			$\delta D = -1.8\%o$
Cul de Sac Plain, Haiti¶	9	springs	$\delta^{18}O = -0.09\%o$
			$\delta D = -0.59\%o$
Martha Brae River basin, Jamaica**	3	precipitation	$\delta D = -0.9$ to $-2.2\%o$
Island of Hawaii, Kilauea volcano area††	50	precipitation, springs	$\delta^{18}O = -0.15$ to $-0.32\%o$
			$\delta D = -0.98$ to $-2.6\%o$

*Gerardo et al. [1993].
†Avis-Isidro et al. [1993].
‡Liu [1984].
§Yasuhara et al. [1995].
‖Stewart and Taylor [1981].
¶Gonfiantini and Simonot [1987].
**Ellins [1992].
††This study.

model for a rising air mass showing that the rate of change of condensate δ values with elevation should increase at high elevations, where condensation occurs as snow rather than rain. Our data (Figure 6b) agree with this model. The high-elevation gradient can be explained by several processes, for example, condensation of snow from a rising air mass or snow melting as it falls and equilibrating to ground temperatures at the collection site.

At earlier stages of this study, before many precipitation data had been obtained for the rain shadow area, the relatively depleted isotopic composition of groundwater south of Kilauea summit was interpreted as an indication that springs at the coast were recharged at elevations above Kilauea summit [Scholl et al., 1992]. As the study progressed it became obvious that precipitation in the rain shadow area was substantially more depleted than in surrounding areas at similar elevations, such that the local precipitation could account for the isotopic composition of the groundwater. The spatial variation in precipitation isotopes observed in this 4200-km² study area indicates that in regions with highly variable microclimates and rainfall conditions, a detailed and (or) carefully designed precipitation sampling network is necessary to avoid misinterpretation.
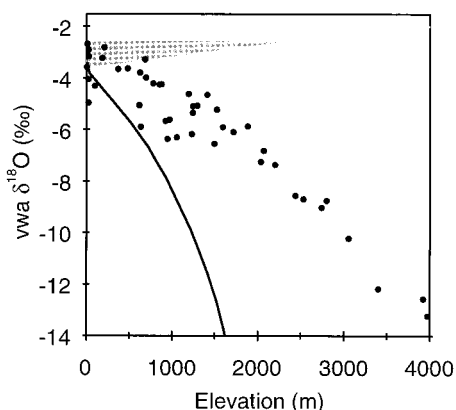
Isotopic Composition of Groundwater

Sample Collection Methods and Average Values

Groundwater samples (Figure 2) were obtained from four sources: (1) wells; (2) groundwater discharge at or just below sea level (coastal springs); (3) cracks or pools in the basalt near the shoreline, where the basal lens is exposed (cracks); and (4) discharge at elevations above 300 m (high-level springs or streams). Most of the wells are water supply wells with pumps installed. Wells were pumped for at least 15 min before sampling or, for those with no pump, sampled from the top of the water column. The wells extend no more than 46 m below sea level (5–45% of the estimated freshwater lens thickness) and are generally uncased or have perforated casing for the entire saturated interval. For the 18 wells sampled in this study there was no consistent correlation between the thickness of the saturated interval and stable isotope composition. The coastal springs and cracks were sampled at low tide, when possible, and samples were generally collected within 45 cm of the surface of the water. Most groundwater sites were sampled three times, in August or September of 1991, 1992, and 1993.

Figure 7. Circular symbols: volume weighted average $\delta^{18}O$ data from Figure 6b. Curve: change in condensate $\delta^{18}O$ composition for a parcel of oceanic vapor ($\delta^{18}O = -13\%o$ [Craig and Gordon, 1965]) that starts condensing at the shoreline and finishes at approximately 1600 m, without input of additional vapor from any source. Shaded triangular area: range of expected $\delta^{18}O$ composition for condensate in equilibrium with oceanic vapor at temperatures between the land surface and maximum altitude of the inversion layer (approximately 2400 m).
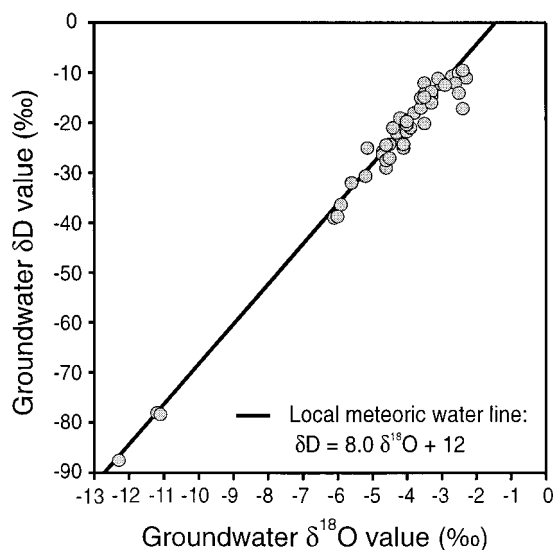
Figure 8. Stable isotope ratios for groundwater, average of samples taken 1991–1993, unadjusted for seawater content.

Samples were analyzed for $\delta^{18}O$, δD, and chemistry, and a subset was analyzed for ^3H. Temperature and pH were measured at time of sampling (see *Janik et al.* [1994] for chemical analyses of the 1991 and 1992 samples). The coastal springs and cracks are brackish, and samples contained varying amounts of seawater. Assuming that the samples were a mixture of seawater and fresh meteoric water, the chloride concentration was used to calculate the isotopic composition of the freshwater end-member, using seawater Cl at 18,900 mg/L, groundwater Cl at 5 mg/L, and seawater at $\delta^{18}O$ and δD of 0‰. Resulting seawater content for the samples ranged from less than 1% to 60%.

The stable isotope composition of water sampled at most springs and wells varied little from year to year, as expected if sampling an extensive groundwater system. The largest variation between yearly samples (as indicated by the standard deviation) was in samples from the often-frozen cracks near the summit of Mauna Loa, probably due to evaporation. Figure 8 shows average groundwater stable isotope ratios, unadjusted for seawater content, for sites sampled in 1991–1993. The samples are generally within the range of -10 to $-40‰$ for δD and -2 to $-6‰$ for $\delta^{18}O$, except for the three samples from cracks near Mauna Loa summit.

Estimation of Recharge Elevations

The $\delta^{18}O$ composition of precipitation was used to estimate the source area of recharge to springs and wells. Interpretation of the recharge areas for groundwater also included some consideration of local geologic structure, rainfall patterns, and sample type. Recharge elevations for each spring and well were estimated in two ways. First, equations (1)–(3) were used to find the elevation at which the measured isotopic content of the groundwater matched the isotopic content of precipitation. The tacit assumption in this case is that the sample represents a single flow path from one area upslope. Second, the isotopic content of the groundwater samples was assumed to represent an integrated sample of all recharge along the slope to some distance above the sampling point. This assumption might not be reasonable for confined groundwater systems but seems plausible for Kilauea's layered basalts, where low-permeability

horizons that restrict recharge are not likely to be areally extensive. Isotopic composition of groundwater at the sampling point was compared to the calculated isotopic composition of recharge from increasing distances upslope, using

$$\delta^{18}O_{sample} = \frac{\sum\limits_{elev=1}^{n} (\delta^{18}O)_n (R)_n}{\sum\limits_{elev=1}^{n} (R)_n} \qquad (4)$$

where $(\delta^{18}O)_n$ is the isotopic value of precipitation for elevation interval n, obtained from (1)–(3), and $(R)_n$ is the estimated recharge volume for the elevation interval n. The intervals were every 152 m of elevation, increasing to every 305 m at higher elevations. Both recharge elevation estimates for each site are shown in Figure 9. For warm springs, recharge elevations were also calculated using δD, in case there had been ^{18}O exchange with rock at some point in the history of the thermal water. For thermal waters where both δD and $\delta^{18}O$ were used to calculate the recharge elevation, the higher elevation is shown. The general model for groundwater flow through most of the study area is that the water table is a subdued replica of the topography, a concept supported by sparse water level data in the Kilauea area [*Takasaki*, 1993]. For each sample site therefore the "flow path" showing the extent of the recharge area is drawn to extend directly upslope from the sampling point, perpendicular to topography, unless geologic and isotopic evidence suggest otherwise.

Recharge $(R)_n$ was estimated as rainfall minus pan evaporation (data from *Ekern and Chang* [1985]) in areas with rainfall greater than 1500 mm/yr. Recharge amounts for drier areas (less than 1500 mm/yr) were based on estimates made for dry areas on Oahu [*Eyre et al.*, 1986; *Takasaki*, 1993] but were arbitrarily increased by 5%, as there is little or no soil developed on parts of Kilauea and Mauna Loa. Recharge in these areas was estimated to be 23% of rainfall for rates of 1300–1500 mm/yr, 22% for 1040–1300 mm/yr, 19% for 790–1040 mm/yr, and 13% for 530–790 mm/yr. These recharge estimates are very approximate; however, the calculated recharge elevations were fairly insensitive to changes of $\pm 10\%$ in the dry-area recharge percentages.

We assumed in this study that rainfall isotopic composition was equivalent to recharge isotopic composition. In the many areas on Kilauea volcano with little soil and high infiltration rates, this is likely a good assumption. In other places, evaporation from soil may tend to isotopically enrich recharge relative to rainfall. An opposite effect may occur on the fringes of the trade wind rainfall areas, where a disproportionate amount of storm rainfall (intense rainfall, isotopically depleted) may recharge the groundwater system, whereas a disproportionate amount of trade wind–type rainfall (drizzle or light rain, isotopically enriched) may be transpired or evaporated. Other sources of error in the recharge elevation estimates include variability on a timescale longer than the 2½ years of the study, uncertainty about whether well samples were a well-mixed representation of the screened interval or flow from one or two discrete fractures, and uncertainty due to the correction for seawater content in coastal spring samples. As a result, we consider the recharge elevations (Figure 9) to be useful mainly for comparative purposes in interpretation of the regional hydrology, rather than absolute values.
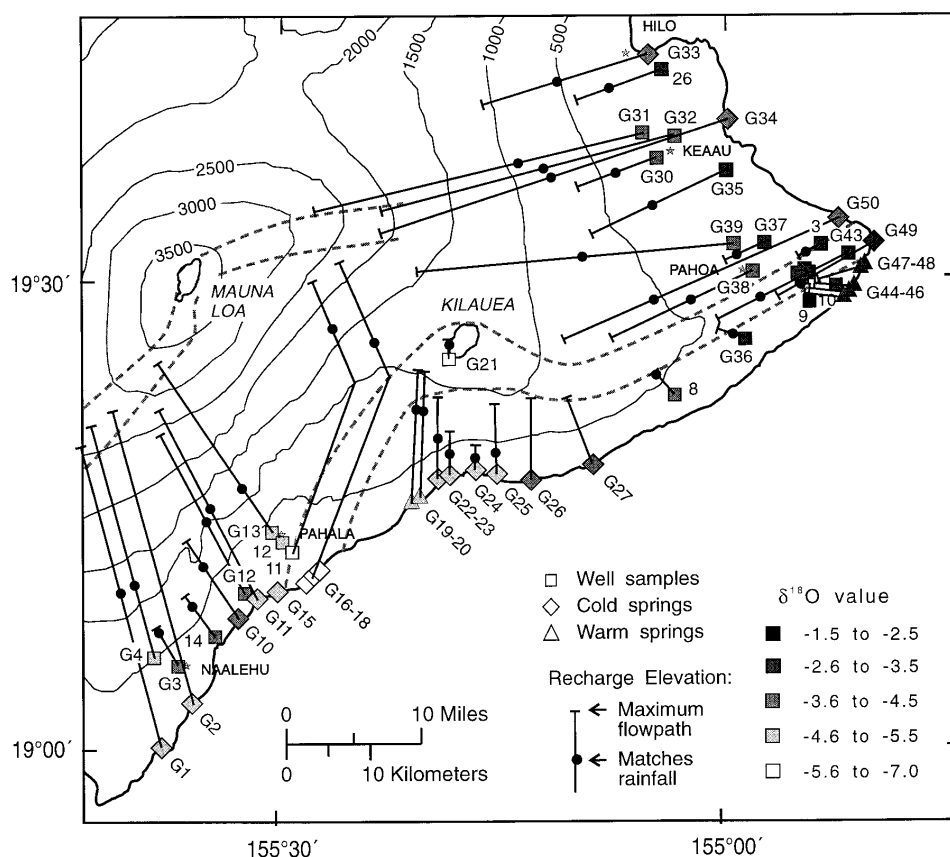
Figure 9. Calculated recharge elevations for groundwater samples. Flow path extending from sample site corresponds to recharge elevation calculated using equation (4); circular symbol on flow path is elevation at which isotopic composition of rainfall matches groundwater measured at site. Stable isotope data for sample sites 14, 11, 12, 8, 9, 10, 3, and 26 are from *McMurtry et al.* [1977].

Tritium

Tritium (^{3}H) was measured in a subset of the groundwater samples to determine their relative age. Tritium in the groundwater samples ranged from 0 to 5.4 TU, and samples were analyzed at the 0.25-TU precision level. We interpreted these data in a very general way, because the ^{3}H levels show only moderate variation, many groundwaters show evidence of mixing with seawater of unknown age, and there are no time series ^{3}H data available for any of the groundwater sample sites. Residence times were estimated by comparing ^{3}H concentrations in the samples to the expected concentrations obtained by solving equations for two simple models of the flow system, the well-mixed reservoir and piston-flow models, using the method of *Yurtsever* [1983]. The well-mixed reservoir model assumes that all inputs to the flow system immediately mix with previous inputs, and concentration in the outflow depends on the rate of decay and residence time. The piston-flow model assumes that inputs to the flow system traverse the flow path without mixing, so the concentration is affected only by radioactive decay along the flowpath. Neither model includes the possibility of mixing between two or more separate reservoirs. Age estimates obtained using these models are approximate but are useful for categorizing the relative ages of water discharging at different localities.

Tritium levels from atmospheric bomb tests, at their maximum in 1963, were about 200 TU in rainfall at Hilo, Hawaii, one tenth of rainfall levels at Ottawa, Canada (IAEA). There-

fore current tritium levels in Hawaii groundwaters are generally lower than in the continental United States. To calculate groundwater ages using the piston-flow and well-mixed models, yearly tritium input to the groundwater system was approximated using the following sources: Yearly volume-weighted average ^{3}H levels in rainfall for 1953–1962 were estimated by linear correlation between the natural logarithm of Ottawa and Hilo levels (IAEA). Tritium levels in rainfall at Hilo were monitored by the IAEA from 1962 to 1969. For 1970–1984, IAEA data from Midway Island were used. Rainfall levels for 1984–1993 were estimated as a logarithmic decrease between the 1984 value and recent rainfall samples from the study area [*Goff et al.*, 1991; *Scholl et al.*, 1995], which had an average ^{3}H content of 2.2 TU. We assumed no in situ production of tritium in the groundwater system and a pre-1953 background rainfall value of 1 TU (R. Michel, U.S. Geological Survey, oral communication, 1992).

Coastal spring samples and well samples were corrected to an apparent freshwater ^{3}H content by assuming a seawater ^{3}H content of 1.6 TU, leading to a conservative age estimate for the ERZ samples, as ^{3}H in seawater in the inland wells was likely much lower. These corrections were not significant for less than 10% seawater. For 10 of 12 sites sampled 2 years in a row, similar residence times were obtained for both samples. Groundwater samples were divided into three categories according to apparent age: (1) Older waters (less than 1.8 TU) which either were recharged more than 35 years ago or were a
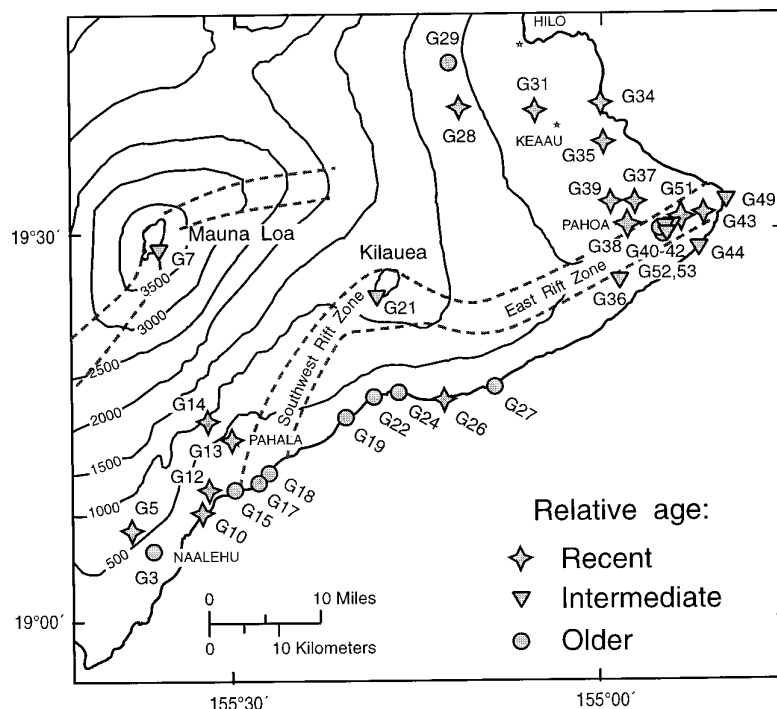
Figure 10. Relative ages for groundwater samples from tritium measurements, estimated using piston-flow and well-mixed reservoir models. "Older" means either older than 35 years or mixed with tritium-dead water; "intermediate" is 18–25 years; and "recent" is 0–17 years.

mixture of recent rainfall and tritium-dead water; (2) intermediate-age waters (greater than 3.5 TU), with an approximately 18–25-year residence time; and (3) recent waters (1.8 to 3.5 TU), with a 0–17-year residence time. Age categories for the groundwater samples based on the well-mixed and piston-flow models are shown on the map in Figure 10. Tritium data for this study are given by *Scholl et al.* [1995].

Interpretation of Groundwater Hydrology

Geologic Framework

The bedrock in the study area is primarily basalt from Kilauea and Mauna Loa, with interspersed ash layers and some buried soils on the flanks of Mauna Loa [*Stearns and Macdonald*, 1946]. The layered structure of the basalt flows creates an anisotropic permeability structure, with near-surface horizontal permeability on the order of 10^{-9} to 10^{-10} m^2 [*Ingebritsen and Scholl*, 1993] and vertical permeability perhaps 10–1000 times less [*Souza and Voss*, 1987]. Hawaiian volcanoes are characterized by rift zones (Figure 1) radiating out from the summit area, along which magma travels to feed eruptions. The rift zones are repeatedly intruded by thin (less than 2 m), nearly vertical, subparallel dikes of dense basalt. The Koae fault zone south of Kilauea Crater (Figure 1) is also thought to be a plane along which dikes are intruded, causing impoundment of groundwater to high levels in the summit area and upper part of both rift zones [*Jackson and Kauahikaua*, 1990].

Occurrence of Groundwater

Previous workers, beginning with *Stearns and Clark* [1930] and *Stearns and Macdonald* [1946], have classified Hawaiian groundwater in terms of three types or occurrences: basal, perched, or dike-confined. Basal groundwater is the freshwater lens in the interior of the island that discharges near sea level at numerous places along the coast of the island [*Stearns and Macdonald*, 1946; *Fischer et al.*, 1966]. This model for groundwater occurrence on Hawaii has recently been updated by *Thomas et al.* [1996], who found freshwater aquifers interlayered with seawater at depths greater than 100 m in a 1-km borehole at the Hilo shoreline. Water discharging at higher elevations in the study area may be perched water, held above the basal lens by ash layers or dense lava flows. High-elevation springs may also be an expression of a locally elevated water table due to some undefined impediment to flow, as evidenced by wells with anomalously high water levels that do not appear to be separated from the basal lens by an unsaturated zone (W. Meyer, U.S. Geological Survey, Honolulu, written communication, 1994). High-elevation springs are common in several areas but are not evenly distributed throughout the region. In the rift zones, dike intrusions can impound groundwater to altitudes substantially above the level of the basal lens. The impounded water may flow parallel to the rift zone, leak out of the rift zone to the basal lens, and (or) be effectively compartmentalized by intersecting dikes. Rift zones may affect the rate and direction of flow of surrounding basal groundwater, depending on the density of dikes in the rift zone and the orientation of the rift zone to regional groundwater flow [*Takasaki and Mink*, 1985].

Compartmentalization by Rift Zones

The rift zones of Mauna Loa and Kilauea control the topography to a large extent. Mauna Loa's rift zones are regional drainage basin divides, but since Kilauea is at lower elevation within the larger drainage basin, we did not assume beforehand that its rift zones subdivided the regional flow system. The isotopic composition of most of the coastal springs to the

north (Hilo-Pahoa area) and the west (Pahala-Naalehu area) of Kilauea's rift zones suggests that the coastal groundwater discharge represents flow from some distance upslope rather than local recharge (Figure 9). However, estimated recharge elevations for groundwater in the area between the rift zones of Kilauea and the ocean suggest that recharge is local to that area. The isotopic evidence indicates that Kilauea's rift zones effectively divide the groundwater system into distinct hydrologic subareas. It had previously been recognized that Kilauea's lower ERZ probably forms a barrier to southward movement of groundwater [*Davis and Yamanaga*, 1968; *Druecker and Fan*, 1976]. The data collected during our study indicate that the area south of the Southwest Rift Zone (SWRZ), Kilauea summit, and the upper ERZ is also hydraulically isolated. This supports *Takasaki*'s [1993] hypothesis that the entire regional flow system around Kilauea is compartmentalized by its rift zones.

The tritium ages also support the idea that the rift zones compartmentalize the regional groundwater system. In general, the older and intermediate-age waters are within or downgradient of the rift zones, while ^{3}H content of all but two of the groundwater samples in areas north and west of the rift zones falls into the recent recharge category (Figure 10). Seawater-corrected ^{3}H content for all but one of the coastal springs south of Kilauea summit is lower than that of recent rainfall, suggesting that the water discharging at the coast is a mixture of recent rainfall and water recharged more than 35 years ago (tritium-dead water). Although rainfall in that area is low compared with most of the island, it is still relatively high (750–1500 mm/yr) and recharge is probably substantial, since the highly permeable young basalts are sparsely vegetated. Travel times for shallow groundwater in that area may not be long enough to account for the presence of tritium-dead water. The high-level, impounded water in the summit area seems a more plausible source for a tritium-dead component in the spring discharge.

Downrift Groundwater Flow

The western boundary between Kilauea and Mauna Loa runs approximately parallel to the trace of the SWRZ. As is the case with the ERZ, gravity measurements indicate that the active rift zone has moved south or seaward over time [*Swanson et al.*, 1976; *Kauahikaua*, 1993], and we assume that there are buried dikes extending some distance north and west of the current surface trace of the rift zone. The lower part of the SWRZ lies downslope from the area of relatively high, isotopically enriched rainfall above Pahala and Naalehu (Figure 3). In contrast, the sparse rainfall over the upper part of the SWRZ has the depleted isotopic composition of the rain shadow area (Figure 6). Four coastal springs (G15–G18) and three Pahala-area wells (11, 12, and G13) (see Figure 9) are considered to be within or influenced by the SWRZ structure [*Scholl et al.*, 1995]. The isotope data from the SWRZ area indicate that (1) recharge from the midslope high-rainfall area is blocked from discharging at the shoreline in the rift zone, and (2) water discharging at the SWRZ springs is recharged above 1500 m. The coastal spring water (G16–G18) is more isotopically depleted than that found in the well near Kilauea summit (G21) and is not heated, suggesting that it enters the upper part of the rift zone from the slopes of Mauna Loa and travels through an older or inactive part of the rift zone. The ERZ of Kilauea is an area of particular hydrologic interest, as geothermal fluids from the lower ERZ are used for electrical power generation and there has been geothermal

exploration in other areas along the rift zone. The entire ERZ is within the trade wind rainfall area, and rainfall is uniform at 3000 mm/yr along most of the length of the rift (Figure 3). There are no wells in the upper ERZ, and wells (e.g., well 8) and coastal springs that once existed below the middle section of the rift have been covered by post-1983 lava flows. Samples obtained for this study were predominantly from coastal warm springs and shallow wells in the lower ERZ (Figure 2).

Apparent recharge elevations for groundwater in the ERZ area are ambiguous. Volume-weighted average precipitation δ^{18}O and δD did not vary significantly between the two collectors at 0 and 370 m elevation along the rift zone (P59, −3.6‰ and −17‰; P51, −3.6‰ and −15‰), although (1a) predicts a 0.6‰ difference in δ^{18}O. Within and south of the lower ERZ, stable isotope composition of groundwater samples was either similar to or more enriched than local rainfall. Unlike the SWRZ, there is no evidence from the shallow wells and springs to support the idea of downrift flow from the vicinity of Kilauea summit, although flow from middle elevations on the rift zone (about 628 m elevation) would have to comprise more than 20% of the sample to be detectable. Isotopic composition of water from deeper wells might show evidence of downrift flow; most of the wells we sampled in this area were less than 40 m below sea level.

Tritium measured in the samples from the ERZ (with two exceptions: G43, approximately 2 m deep, and G51) indicated intermediate to old waters. Tritium content in most shallow wells and coastal springs in the lower ERZ was significantly higher than in groundwater to the north of the ERZ and in rainfall. Tritium in the deep geothermal wells (G52 and G53) was very low. Thus water in and south of the lower ERZ has a longer residence time than waters north of the rift. Since the freshwater stable isotopes do not indicate downrift flow of groundwater, low-permeability dikes in the rift may contribute to longer residence time by impeding outflow. Thermally driven convective circulation within and near the rift zone may also contribute older water from depth to the springs and shallow wells.

Thermal Waters

Groundwater sampled in the area south of Kilauea's rift zones and summit is warmer than groundwater sampled in other parts of the study area [*McMurtry et al.*, 1977; *Takasaki*, 1993; this study]. Average temperatures measured during this study in coastal springs downgradient of the summit and rift zones ranged from 24° to 39°C, whereas coastal spring temperatures west and north of the rift zones ranged from 18° to 22°C. An explanation for the spring chemistry south of the lower ERZ is that seawater in saturated rock below the freshwater lens is heated to about 165°C, boils to 100°C and loses steam, and then mixes with the overlying freshwater lens and flows toward the shoreline, where it is further mixed with seawater from the ocean as it discharges [*Janik et al.*, 1994]. The shallow monitoring wells (G40–G42) and many of the warm springs in the lower ERZ had δD and δ^{18}O values that after correcting for seawater content, were more enriched than local rainfall. This might be expected for δ^{18}O, if there was some exchange with rock, but not for δD. Some possible explanations are the following: (1) boiling of the near-rift seawater end-member resulted in an isotopic composition more enriched than the seawater value (δD and δ^{18}O = 0‰) which we used as the correction; (2) data were not sufficient to define a long-term rainfall average for the area; and (3) there was vapor loss from the heated water through the unsaturated zone along the flow path. Recharge of the freshwater component for the warm

springs and wells was assumed to be local (Figure 9), but factors other than elevation of recharge appear to be influencing isotopic composition in those samples.

Stable isotopes in the coastal springs south of Kilauea summit (G22–G25) are also enriched relative to upslope rainfall. Isotopic composition of rainfall over Kilauea summit (the trade wind/rain shadow boundary) is extremely variable, with $\delta^{18}O$, δD varying from $(-4.6‰, -22‰)$ to $(-6.3‰, -38‰)$ over a small (15 km^2) area (sites P26, P28, and P24; Figure 2). Stable isotopes in well G21 near the summit $(-5.6‰, -32‰)$ are close to the average of the two nearest rain collectors, and somewhat more isotopically enriched than precipitation falling downslope in the rain shadow area (average $\delta^{18}O$, δD of $-6.0‰, -37‰$ from P20–P22). Isotopic composition of rainfall measured in collectors at the coast ranged from $(-4.1‰, -25‰)$ at P23, in the east, to $(-5.0‰, -31‰)$ at P10, in the west.

This variation complicates interpretation of the groundwater system, because (1) the coastal groundwater samples are more isotopically enriched than the average rainfall immediately upslope and are most similar to either coastal rainfall or Kilauea summit rainfall, and (2) the ^3H data and elevated temperature suggest that the coastal springs south of Kilauea summit contain a component of impounded groundwater from the summit area. Perhaps there is some degree of thermal stratification, where warm water leaking from the summit area floats on the top of the freshwater lens, as was noted by *McMurtry et al.* [1977] for wells in the lower ERZ. Despite tidal mixing near the coast, some thermal stratification may be preserved, so that the samples taken at sea level in the cracks and springs contain a larger portion of thermal water.

Summary

Significant areal differences in precipitation isotopes on the Island of Hawaii correlate strongly with general climatological patterns. High-rainfall areas have frequent rains owing to the orographic lifting of moist air carried by the trade winds or by a thermally driven seabreeze cycle. The isotopic content of cumulative samples from these high-rainfall areas agrees well with that predicted from first-stage condensation of atmospheric vapor above the ocean. Precipitation isotopes get more depleted with increasing elevation and distance inland, as expected, owing to decreasing temperature and "rain out" effects. In areas of lower rainfall, most rainfall occurs during storms. In the rain shadow of Kilauea volcano, isotopic content of precipitation samples also varies with elevation and distance inland but is more depleted than at the same elevations in the trade wind areas. The depleted isotopic composition of precipitation in the rain shadow can be explained by the processes affecting rainfall in storms. Although the original source of vapor in storms is the oceanic atmosphere, condensation temperatures for large storms are generally lower, and such storms are often producing rain before they reach the island. For the high-elevation area above the orographic rainfall influence, precipitation isotopes also vary with elevation, but the rate of decrease in δ values with elevation is greater than for lower-elevation areas, presumably owing to the difference in isotopic fractionation factors when condensation temperatures are below 0 to $-15°C$. It seems reasonable to expect that similar relations between precipitation isotope content and elevation may apply to other islands and possibly coastal continental areas at similar latitudes. In particular, the other Hawaiian Islands have very similar temperature and climatological patterns. It may be feasible to use stable isotopes as tracers for groundwater flow on the other islands with fewer precipitation samples, just a sufficient number to establish that gradients are similar. However, this study also underscores the importance of a collection strategy that takes areal variations in rainfall patterns into account, as precipitation isotopes vary significantly with microclimates.

Stable isotopes worked well as tracers of groundwater flow paths in the southeast part of the Island of Hawaii. Tritium was used to distinguish broad age categories for groundwater in the study area. These isotope data establish that Kilauea's rift zones effectively compartmentalize the regional groundwater system, with relatively local recharge and longer residence times for groundwater in and downgradient of the rift. The isotope data also show that the impounded water near Kilauea summit is recharged locally. Heated groundwater from the Kilauea summit area may leak southward to the springs at the shoreline. Both stable isotope and tritium data indicate that part of Kilauea's SWRZ acts as a conduit for groundwater flow. Stable isotopes from the shallow wells and thermal springs in the lower ERZ showed no evidence for downrift flow. Therefore elevated ^3H levels in the lower ERZ shallow groundwater suggest low permeabilities rather than a long flow path.

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